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EFFECT OF CLOUD-PRECIPITATION-OCEAN MIXED LAYER FEEDBACK ON DRAG COEFFICIENT

Peter C. Chu and Roland W. Garwood, Jr.

Department of Oceanography
Naval Postgraduate School
Monterey, CA 93943, USA

1. INTRODUCTION

From similarity theory, the drag and thermal exchange coefficients (C_D, C_H, and C_E) for 10m above the ocean surface depend on the atmospheric stability parameter (z/L_a) where L_a the Obukhov length scale for the atmospheric surface layer. Cloud, precipitation, and ocean mixed layer (OML) thermodynamical feedback affects the atmospheric stability parameter in two ways. Clouds reduce the incoming solar radiation at the ocean surface by scattering and absorption, which cools (relatively) the ocean mixed layer. The OML cooling lowers the sensible and latent heat fluxes from the ocean surface, which increases the atmospheric stability parameter. On the other hand, precipitation from the clouds dilutes the surface salinity, stabilizing the upper ocean, and reducing OML deepening. The ocean mixed layer may be caused to shallow if the downward surface buoyancy flux is sufficiently enhanced by the precipitation. When the ocean surface receives a downward net surface heat flux, the reduction in the OML depth will increase the sea surface temperature (SST) by concentrating the net radiation plus heat fluxed downward across the sea surface into a thinner layer (Fig.1). The increase of SST augments the sensible and latent heat fluxes from the ocean surface, which decreases the atmospheric stability parameter. On the other hand, when the ocean surface has a net surface heat loss, the reduction in the OML depth will decrease SST due to the decrease of the thermal inertia. The decrease of SST reduces the sensible and latent heat fluxes from the ocean surface, which increases the atmospheric stability parameter. The variation of atmospheric stability parameter in turn changes the drag and thermal exchange coefficients. In this study, a simple coupled cloud-precipitation-ocean mixed layer model proposed by Chu et al. (1990) and Chu and Garwood (1991) is used to compute the time rate change of drag and thermal exchange coefficients caused by this mechanism.

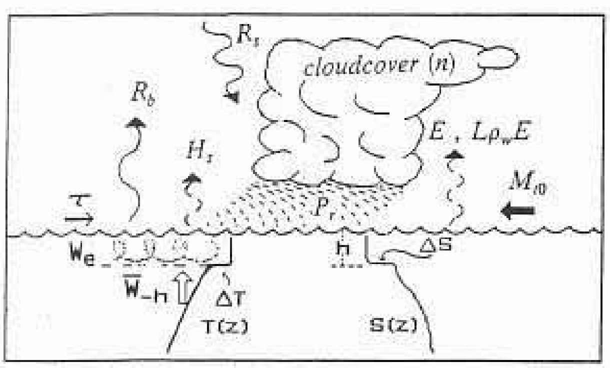

Fig.1. Main physical processes in the two adjacent boundary layers (from Chu et al., 1990).

2. DRAG AND THERMAL EXCHANGE COEFFICIENTS

From the Monin-Obukhov similarity theory, the momentum, heat, and moisture fluxes from the air-ocean interface are directly related to the mean profiles of wind, temperature, and mixing ratio:

$$\frac{\partial \overline{U}_a}{\partial z} = \frac{u_*}{\kappa z}\phi_M\left(\frac{z}{L_a}\right), \quad \frac{\partial \overline{T}_a}{\partial z} = \frac{T_*}{z}\phi_H\left(\frac{z}{L_a}\right),$$

$$\frac{\partial \overline{q}}{\partial z} = \frac{q_*}{z}\phi_E\left(\frac{z}{L_a}\right) \tag{1}$$

where ϕ_M, ϕ_H, ϕ_E are similarity functions; u_* is the atmospheric friction velocity; κ is von Karman constant; T_* and q_*, the atmospheric surface layer temperature and humidity scales, represent the sensible and latent heat fluxes across the ocean surface (upward positive),

$$Q_S \equiv -\rho_a c_{pa} \kappa u_* T_*, \quad Q_E \equiv -\rho_a L_v \kappa u_* q_* \tag{2}$$

where ρ_a, c_{pa}, L_v are surface air density (characteristic value), specific heat of air, and latent heat of vaporization. For a near-neutral atmospheric surface layer (small z/L_a), the drag coefficient is computed by

$$C_D = \frac{\kappa^2}{\left[\ln\left(\frac{z_{10}}{z_0}\right) + b\frac{z_{10}}{L_a}\right]^2} \tag{3}$$

where $z_{10} = 10m$, z_0 is the roughness length, and

$$b \equiv \left[\frac{\partial \phi_M}{\partial(z/L_a)}\right]_{z/L_a=0} \simeq 3.0$$

from the Kansas experiment (Businger et al., 1971). The atmospheric Obukhov length scale L_a is defined by

$$L_a \equiv -\frac{\overline{T}_{va} u_*^3}{\kappa g\left(\frac{Q_S}{\rho_a c_{pa}} + 0.61\overline{T}_a\frac{Q_E}{\rho_a L_v}\right)} \tag{4}$$

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where $\overline{T}_a$ and $\overline{T}_{va}$ are mean air and virtual temperatures.

In the neutral atmosphere, $z/L_a = 0$, the drag coefficient becomes

$$C_{DN} = \kappa^2[\ln(\frac{z_{10}}{z_0})]^2 \qquad (5)$$

If bz_{10}/L_a is small compared to $\ln(z_{10}/z_0)$, the drag coefficient for a non-neutral atmosphere is related to the drag coefficient for a neutral atmosphere (C_{DN}) by

$$C_D = C_{DN}(1 - \frac{2bz_{10}/L_a}{\ln(z_{10}/z_0)}) \qquad (6)$$

where all parameters on the righthand-side except L_a are independent of surface heat and moisture fluxes. Therefore, the time rate of change of drag coefficient due to the change of surface heat and moisture fluxes is influenced only by the change of L_a:

$$\frac{1}{C_{DN}}\frac{\partial C_D}{\partial t} = \frac{2bz_{10}/L_a^2}{\ln(z_{10}/z_0)}\frac{\partial L_a}{\partial t} \qquad (7)$$

Substitution of (4) into (7) leads to a expression representing the effect of change of surface heat and moisture fluxes on the time rate change of drag coefficient:

$$\frac{1}{C_{DN}}\frac{\partial C_D}{\partial t} = \frac{2bz_{10}\kappa g}{\ln(z_{10}/z_0)\overline{T}_{va}u_*^3}[\frac{1}{\rho_a c_{pa}}\frac{\partial Q_S}{\partial t} + \frac{0.61\overline{T}_a}{\rho_a L_v}\frac{\partial Q_E}{\partial t}] \qquad (8)$$

The bulk formulas for the sensible and latent heat fluxes are

$$Q_S = \rho_a c_{pa} C_H (T_s - T_{10})\overline{U}_a, \quad Q_E = \rho_a L_v C_E [q_s(T_s) - q_{10}]\overline{U}_a \qquad (9)$$

where C_H and C_E are the 10m thermal exchange coefficients. If the mean surface wind and air temperature is determined mostly by the large scale atmospheric motion, substitution of (9) into (8) leads to

$$\frac{1}{C_{DN}}\frac{\partial C_D}{\partial t} = M\frac{\partial T'_s}{\partial t} \qquad (10)$$

where

$$M \equiv \frac{2bgz_{10}}{\lambda \overline{T}_{va}\overline{U}_{10}}(1 + 0.61\overline{T}_{10}\frac{dq_s}{dT_s}), \quad \lambda \equiv \frac{\int \phi_H(0) d\ln z}{\int \phi_M(0) d\ln z} \simeq 0.74$$

Here we assume that

$$\frac{1}{C_{DN}}\frac{\partial C_D}{\partial t} \simeq \frac{1}{C_H}\frac{\partial C_H}{\partial t} \simeq \frac{1}{C_E}\frac{\partial C_E}{\partial t} \qquad (11)$$

If we use $\overline{T}_{va} \simeq \overline{T}_{10} \simeq \overline{T}_s \simeq 288°K$, and $\overline{U}_{10} \simeq 5m/s$ the M-values is roughly estimated as $M \simeq 0.13K^{-1}$.

3. OCEAN MIXED LAYER DYNAMICS

Most models that include thermodynamic effects include the assumption that the upper ocean layer as a well-mixed turbulent boundary layer which exchanges heat and moisture with the atmosphere and entrains water from deeper layer. The heat and salinity equations take the forms

$$h_w\frac{\partial T_s}{\partial t} = -w_e(T_s - T_{-h}) - \frac{F}{\rho_w c_{pw}} + A_T \qquad (12a)$$

$$h_w\frac{\partial S}{\partial t} = -w_e(S - S_{-h}) + (E - Pr)S + A_S \qquad (12b)$$

where h_w is the mixed layer depth; T_{-h} and S_{-h} are temperature and salinity at the base of the entrainment zone; E is the surface evaporation rate; Pr is the precipitation rate; A_T and A_S are horizontal advection for temperature and salinity, respectively; and

$$F \equiv R_b - R_s + Q_S + Q_E \qquad (13)$$

is the net heat flux at the air-ocean interface, where R_s and R_b are solar and back longwave radiation at the ocean surface. The entrainment velocity w_e is parameterized by Chu and Garwood (1988) as

$$w_e = \frac{\Lambda_w(P)}{gh_w[\alpha(T_s - T_{-h}) - \beta(S - S_{-h})]} \qquad (14)$$

where

$$P \equiv (C_1 u_{w*}^3 - C_2 B h_w)$$

is the surface forcing function; C_1 and C_2 are tuning coefficients; $u_{w*} = (\rho_a/\rho_w)^{1/2}u_{a*}$ is the water surface friction velocity; and

$$B \equiv \frac{\alpha g F}{\rho_w c_{pw}} + \beta g(Pr - E)S \qquad (15)$$

is the downward ocean surface buoyancy flux. The symbol Λ_w is a Heaviside function of P. When $P > 0$, there is sufficient turbulent kinetic energy to entrain and mix water from below and $\Lambda_w = 1$, which represents the entrainment regime. The time rate change of mixed layer depth is

$$\frac{\partial h_w}{\partial t} = w_e - w_{-h} \qquad (16a)$$

Here w_{-h} is the mean vertical velocity at the mixed layer base. When $P \le 0$, there is insufficient turbulent kinetic energy to entrain water from below, and w_e is set to be zero, i.e., $\Lambda_w = 0$. This is called the shallowing regime. The mixed layer depth is calculated from $P = 0$, and it is directly proportional to the oceanic Obukhov length scale, L_w:

$$h_w = L_w = \frac{C_1 u_{w*}^3}{C_2 B} \qquad (16b)$$

Chu and Garwood (1991) discussed the effect of cloud amount and precipitation perturbations (n', Pr') on the OML dynamics. Three parameters were found important:

$$\Gamma \equiv \frac{1}{\rho_w c_{pw}\bar{h}_w}\frac{\partial(R_b - R_s)}{\partial n},\quad \xi \equiv \frac{\partial \overline{Pr}}{\partial n},\quad \gamma \equiv \frac{\beta g \bar{S}(\overline{Pr}-\bar{E})}{\bar{B}} \qquad (17)$$

where Γ is the cloud-radiative forcing at the ocean surface; ξ is the cloud-precipitation parameter, representing precipitation brought on by unit cloud amount; and γ is the surface water budget index representing the relative importance of surface freshwater influx in the surface buoyancy flux.

Let us consider the effect of cloud and precipitation perturbation on SST for the two different OML regimes. After mathematical manipulation, the SST perturbation caused by cloudiness disturbance for the entrainment regime is

$$\frac{\partial}{\partial t}\left(\frac{\partial}{\partial t}+\tau_T^{-1}\right)T'_s = -(1-C_2)\Gamma\frac{\partial n'}{\partial t} + C_2\tau_c^{-1}\left(\Gamma - \frac{\beta\bar{S}}{\alpha\bar{h}_w}\xi\right)n' \qquad (18a)$$

and for the shallowing regime is

$$\frac{\partial T'_s}{\partial t} = \left[-\gamma\Gamma + (1-\gamma)\frac{\beta\bar{S}}{\alpha\bar{h}_w}\xi\right]n' \qquad (18b)$$

where

$$\tau_T^{-1} \equiv \frac{1}{\rho_w c_{pw}\bar{h}_w}\frac{\partial\bar{F}}{\partial T_s} \simeq \frac{1}{2\,year}$$

$$\tau_c^{-1} \equiv \frac{\bar{F}}{\rho_w c_{pw}\bar{h}_w\Delta T} \simeq \frac{1}{18\ day}$$

for $\bar{F}\simeq 100\,Wm^{-2}$, $\bar{h}_w \simeq 50\,m$, and $\Delta T \simeq 3^\circ K$. Here, τ_T is the residence time for SST perturbation, and τ_c is the time needed for OML warming up to the deeper water temperature by $\bar{F}$.

4. CHANGE OF DRAG COEFFICIENT DUE TO HEAVY PRECIPITATION

Fig.2 shows the feedback loop among cloud, precipitation, OML, and the drag and thermal exchange coefficients (C_D, C_E, C_H). Cloud and precipitation changes the surface buoyancy flux, which in turn changes SST through the variation of both surface net heat flux and the OML depth. The SST variation changes the surface sensible and latent heat fluxes, which causes the modification of the drag and thermal exchange coefficients.

For a very heavy transient precipitation, such as $Pr > 0.1m/day$,

$$\left|\frac{\beta\bar{S}}{\alpha\bar{h}_w}\xi\right| \gg |\Gamma|$$

when the time scale for the precipitation is much shorter than τ_T, the SST equation for the entrainment regime becomes

$$\frac{\partial^2 T'_s}{\partial t^2} \simeq -(1-C_2)\Gamma\frac{\partial n'}{\partial t} - C_2\tau_c^{-1}\frac{\beta\bar{S}}{\alpha\bar{h}_w}\xi n' \qquad (19a)$$

the SST equation for the shallowing regime is

$$\frac{\partial T'_s}{\partial t} \simeq (1-\gamma)\frac{\beta\bar{S}}{\alpha\bar{h}_w}\xi n' \qquad (19b)$$

Substitution of (19a,b) into (10) leads to

$$\frac{1}{C_{DN}}\frac{\partial^2 C_D}{\partial t^2} \simeq -M\left[(1-C_2)\Gamma\frac{\partial n'}{\partial t} + C_2\tau_c^{-1}\frac{\beta\bar{S}}{\alpha\bar{h}_w}\xi n'\right] \qquad (20a)$$

for the entrainment regime, and

$$\frac{1}{C_{DN}}\frac{\partial C_D}{\partial t} \simeq M(1-\gamma)\frac{\beta\bar{S}}{\alpha\bar{h}_w}\xi n' \qquad (20b)$$

for the shallowing regime.

In determining the change of the drag coefficient, two parameters ξ, γ are crucial when OML is in the shallowing regime; however, ξ, Γ are important when OML is in entrainment regime. In the following, we only present the results for the shallowing regime.

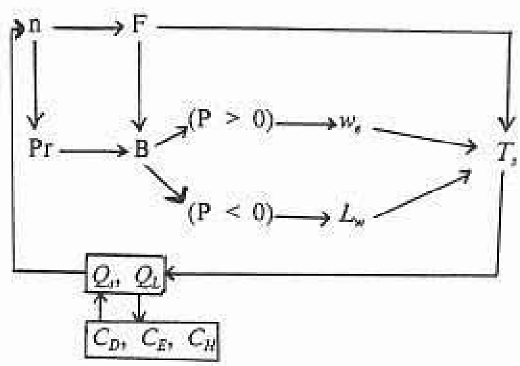

Fig.2. Feedback loop among cloud, precipitation, OML, and the drag and thermal exchange coefficients.

Fig.3 shows the dependence of $1/C_{DN}\partial C_D/\partial t$ on the cloudiness disturbance n' for different values of γ (0.1, 0.5, 0.9, 2.5, 5.0) at $\xi = 0.5\,m/day$. For a heavy precipitation case, i.e., $Pr - E \gg 0$, we have $B > 0$. From (17), $\gamma > 1$, represents the upward net heat flux at the ocean surface (ocean losing heat at the surface); and $\gamma < 1$, represents the downward net heat flux at the ocean surface (ocean gaining heat at the surface), as shown in Fig.4. Two distinct effects of cloudiness disturbance on the time rate change of drag coefficient were found depending on different values of γ. For $\gamma < 1$ (ocean surface losing heat), the atmospheric surface layer is destabilized and the drag coefficient is increased; however, for $\gamma > 1$ (ocean surface gaining heat), the atmospheric surface layer is stabilized and the drag coefficient is decreased. For $\gamma = 5.0$, and $n' \simeq 0.5$,, $1/C_{DN}\partial C_D/\partial t \simeq -0.36/day$, which means that the drag coefficient C_D will be reduced by 36% during 1 day.

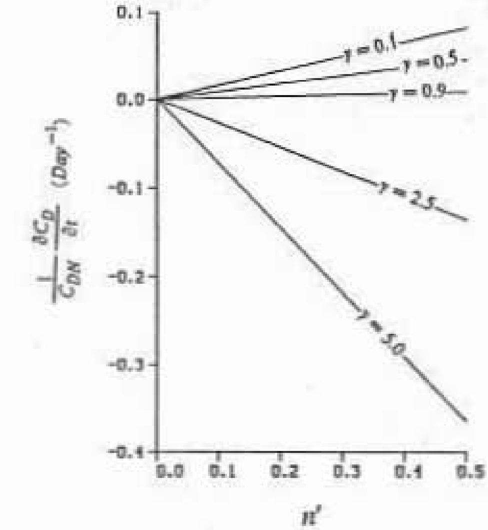

Fig.3. Dependence of $1/C_{DN}\partial C_D/\partial t$ on the cloudiness disturbance n' for different values of γ (0.1, 0.5, 0.9, 2.5, 5.0).

The dependence of the time rate change of the drag coefficient on n' for five different values of ξ is shown in Fig.4 for $\gamma = 0.1$ and in Fig.5 for $\gamma = 5.0$. The cloud-precipitation-OML effect on the drag coefficient strengthens as ξ increases. If the precipitation rate from a severe storm is 0.125 m/day, and during the storm the cloudiness only changes 25%, the cloud-precipitation coupling coefficient $\xi \simeq 0.5 m/day$. For $\gamma = 0.1$ (ocean surface losing heat) and $\xi \simeq 0.5 m/day$, $1/C_{DN}\partial C_D/\partial t \simeq 0.01 - 0.1/day$, which means that the drag coefficient increases 1-10% during 1 day (Fig.4). For $\gamma = 5.0$ (ocean surface gaining heat) and $\xi \simeq 0.5 m/day$, $1/C_{DN}\partial C_D/\partial t \simeq -0.2/day$, for $n' \simeq 0.25$, which means that the drag coefficient decreases around 20% during 1 day (Fig.5).

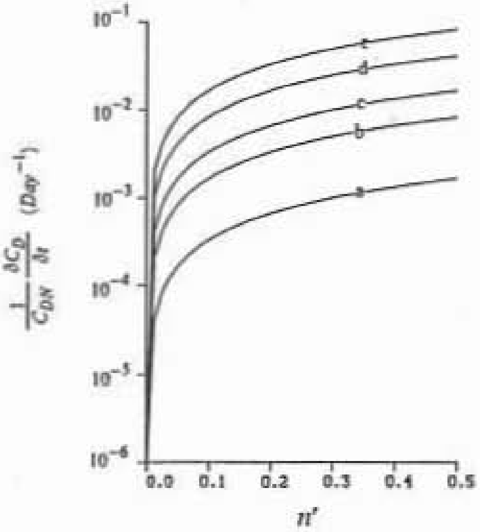

Fig.4. Dependence of $1/C_{DN}\partial C_D/\partial t$ on the cloudiness disturbance n' for $\gamma = 0.1$ (ocean losing heat), and for: (a) $\xi = 0.01 m/day$, (b) $\xi = 0.05 m/day$, (c) $\xi = 0.1 m/day$, (d) $0.25 m/day$, and (e) $0.5 m/day$.

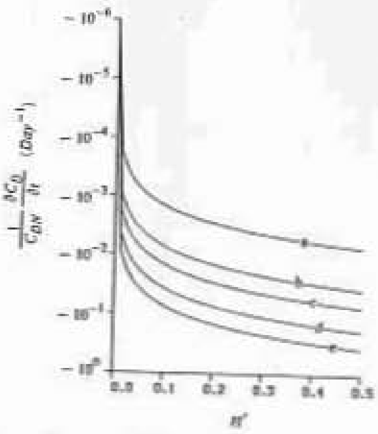

Fig.5. Dependence of $1/C_{DN}\partial C_D/\partial t$ on the cloudiness disturbance n' for $\gamma = 5.0$ (ocean gaining heat), and for: (a) $\xi = 0.01 m/day$, (b) $\xi = 0.05 m/day$, (c) $\xi = 0.1 m/day$, (d) $0.25 m/day$, and (e) $0.5 m/day$.

5. DISCUSSION

(a) In this study, a new thermodynamic feedback mechanism among cloud, precipitation, and OML is found important for determining the drag coefficient. This process should be considered supplemental to the usually dominating factors of surface roughness on the neutral drag coefficient.

(b) Heavy precipitation affects the drag coefficient greatest, especially for a warm ocean surface (i.e., ocean losing heat the atmosphere). The drag coefficient can increase 20-30% within 1 day.

(c) It is not the purpose of this first study to parameterize the drag coefficient in detail. In the future, we will use more sophisticated large eddy simulation (LES) model for both atmosphere and ocean to do further research.

(d) Similar to the drag coefficient, the thermal exchange coeffcients C_H and C_E are also affected by this cloud-precipitation- OML feedback mechanism.

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